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Water Balance at the Southern Limit of the Californian Mixed-Conifer Forest and Implications for Extreme-Deficit Watersheds

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This study estimates the soil water balance at forest/chaparral, upper forest, and wet-meadow sites in the Sierra San Pedro Mártir (SSPM), Baja California, Mexico, and compares the results to previously published data for similar sites in neighboring California, USA. Changes in soil water storage (0–80 cm) were determined by using a neutron probe, deep drainage (> 80 cm) was measured in

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minilysimeters, and runoff was estimated in 1 m^2 plots. Average annual precipitation was 714 ± 226, 674 ± 137, and 551 ± 112 mm at the lower forest, upper forest, and wet-meadow sites, respectively. Evapotranspiration (ET) was 390 ± 122, 478 ± 87, and 446 ± 122 mm, or 55%, 71% and 81% of precipitation. The relationship between average precipitation and average ET for SSPM was similar to published long-term ET data for extreme-deficit watersheds in southern California. Regression analyses of streamflow and ET data in review studies revealed that, as water deficit increases, the relationship between precipitation and ET becomes increasingly stronger, while that between precipitation and streamflow becomes concomitantly weaker. The slope of the regression line gives the proportion of precipitation lost to streamflow or ET, while the intercept indicates the amount of soil water recharge required annually before streamflow can occur. Available data from long-term studies in extreme-deficit watersheds revealed a near-zero intercept, and imply that 69% of precipitation is evapotranspired from the surface soil, Also, 15% is apparently transpired by deep roots extracting water below the solum, and the remaining 16% goes to streamflow.

Keywords Baja California, chaparral, evapotranspiration, Mediterranean climate, soil moisture, streamflow, water budget

Consumptive use of water by vegetation accounts for most of the precipitation that falls on forested watersheds, constraining water yield to lowland streams. The accelerating agricultural and urban development of arid northern Baja California, already densely populated and dependent on transmontane Colorado River water, is increasing the pressure to use streamflows from mountain catchments. The Peninsular Range of the Californias spans the economically dynamic Mexico-USA border region and is the southeastern limit of North America's Mediterranean-type region. This climate type, characterized by cool-wet winters and warm-dry summers, poses a unique dilemma for transpiration by plants because the availability of water and energy are out of phase (Turner, 1991). The water balance at the southern limit of the Californian mixed-conifer forest is of ecological interest, moreover, because the seasonal timing of water and energy supplies, especially of actual (or "standard") ET and water deficits, is strongly correlated with the distributions of plant formations in North America (Stephenson, 1998).

Many published studies of both streamflow and ET in summer-dry mountain watersheds were done in neighboring southern California (e.g., Shachori & Michaeli, 1965; Rutter, 1968; Arkley, 1981; Bosch & Hewlett, 1982; Turner, 1986). However, little information is available on Baja California's Sierra San Pedro Mártir (SSPM), which is the southern terminus of the Peninsular Range and home to the southernmost Californian mixed-conifer forest (Minnich et al., 1997).

Although Shachori and Michaeli (1965) found a very strong relationship (R^2 = 0.93) between precipitation and streamflow in data from the USA southwest, Europe, and Israel, and Turner (1986, 1991) reported a strong correlation (R^2 = 0.76) between precipitation and ET (calculated as precipitation minus streamflow) in 68 California watersheds, little has been done to analyze these relationships in worldwide data sets, or to compare the results of ET and streamflow studies. The aims of this study were (1) to determine the average annual water balance at two forest and one wet-meadow sites in the Sierra San Pedro Mártir, (2) to compare those results with published data for similar sites in neighboring southern California, and (3) to characterize the relationships between precipitation, ET, and streamflow in these ecosystems.

Materials and Methods

Three study sites were established in the Sierra San Pedro Mãrtir (Figure 1, Table 1), two in the conifer forest on soils derived from granitic rocks, and a third in a

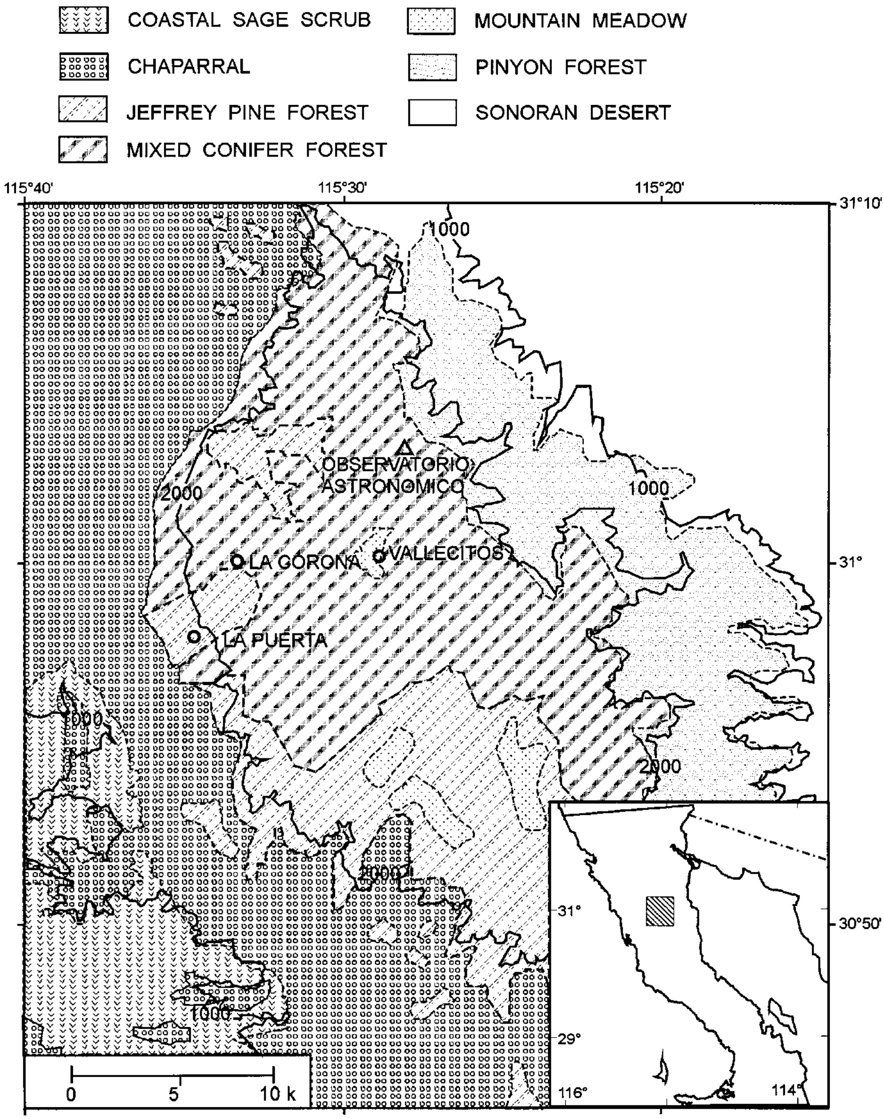

FIGURE 1 Topographic and vegetation map of the Sierra San Pedro Mártir showing the location of the study sites. Elevations are in meters above sea level. Dominant plants of the coastal sage scrub are: *Artemisia californica* Less, *Eriogonum fasciculatum* Benth., and *Encelia californica* Nutt.; of the chaparral, *Adenostoma fasciculatum* Hook & Arn., *A. sparsifolium* Torr., and *Ceanothus greggii perplexans* (Trel.) Jepson; of the Jeffrey pine forest, *Pinus jeffreyi* Grev. & Balf., *Arctostaphyllos peninsularis* Wells, and *Arctostaphyllos pringlei* C. Parry; of the mixed-conifer forest: *P. jefrreyi*, *Abies concolor* Lindl., and *P. lambertiana* Dougl.; of the mountain meadows: *Juncus mexicanus* Willd., *Carex douglasii* Boott., *Achillea millefolium* (Poll.) Jeps., and *Potentilla wheeleri* S. Wats.; of the pinyon forest: *Pinus monophylla* Torr. & Frem., *P. quadrifolia* Parl., and *Juniperus californica* Carr.; of the Sonoran Desert: microphyllous woodland (arroyos), *Acacia greggii* A. Gray, *Cercidium floridum* Benth. ex A. Gray, *Prosopis glandulosa* Torr.; Creosote Bush scrub, *Larrea tridentata* Sessé & Moc. ex DC, *Agave deserti* Engelm., *Ambrosia dumosa* (A. Gray) Payne, and *Encelia farinosa* A. Gray.

TABLE 1 Characteristics of the Study Sites

Study Site (place name)	Altitude (m)	Mean annual precipitation[a] (mm)	Physiographic position	Vegetation	Cover (%) Trees	Cover (%) Shrubs	Cover (%) Herbaceous	Soil classification[b] depth (cm)
Lower Forest (La Puerta)	1,980	699	Upper slope, western escarpment	Jeffrey-pine forest-chaparral boundary	43	16	4	Typic Xeropsamments 58
Upper Forest (La Corona)	2,470	700	Just east of crest, western escarpment	Mixed-conifer forest	25	6	5	Typic Xeropsamments 60
Mt. Meadow (Vallecitos)	2,380	552	Central upper plateau	Wet meadow	0	0	53	Aeric Haplaquepts >150

[a]Minnich et al., 2000.
[b]R.C. Graham, University of California, Riverside, personal communication.

mountain meadow surrounded by "islands" of mixed-conifer forest on soils derived from granitic alluvium. The water balance was estimated indirectly from the hydrological equation as described by Rutter (1968), $ET = P \pm \Delta S - R - D$, where ET is evapotranspiration, P is precipitation, ΔS is the change in soil water storage, and D and R are drainage and runoff from the system. Precipitation was measured in a standard bulk rain gauge at each site, and changes in soil water storage were measured by using a neutron probe (Hydroprobe Moisture Depth Gauge, Model 503, Campbell Pacific Nuclear Corp., Martinez, California, USA) in 5-cm diameter PVC access tubes. The snowpack was measured by using a standard USDA Soil Conservation Service snow probe. Runoff was measured in two each 1 m^2 plots installed in 1991 at the two forest sites, and a relationship was derived between monthly precipitation and runoff which was applied to the earlier forest site data and for the mountain meadow. These small 1 m^2 runoff plots were used in order to prevent the intense precipitation of summer thunderstorms from overwhelming the capacity of the runoff capture and storage units, which were unattended, and also because of the fast infiltration in the coarse granitic sand during gentle winter rains. Deep drainage ($>$ 80 cm) was measured in minilysimeters constructed from a 200-L plastic barrel to which was attached (with Silicone® sealant) a reservoir made from the bottom half of a 60-L plastic barrel. The top of the large barrel was removed and drainage into the reservoir below was permitted by numerous holes about 2-mm diameter drilled into the bottom of the upper barrel.

The large upper barrel was back-filled with the same soil dug from the pit, and packed as close to the density of the natural soil as possible. At the forest sites, rock contact was at about 60 cm and a hollow was excavated in the rock for the 30-L reservoir. At each site, two minilysimeters were installed about 10 m apart, and two neutron-probe access tubes were placed, each about 10 m from a minilysimeter. A neutron probe access tube was also installed in the center of the soil-filled upper barrel in each minilysimeter. To measure deep drainage, water was removed from the reservoir by using a suction pump through a length of 1-cm diameter plastic tubing. The minilysimeters worked adequately in the forest sites, but the mountain meadow had at times several cm of flowing surface water during the wet winters of 1990–1991 and 1991–1992. To estimate deep drainage in the meadow, the relationship between precipitation and deep drainage obtained during the drier periods was used to estimate drainage during the period of flooding.

For all calculations, data from the two neutron-probe access tubes inside and outside the minilysimeters were averaged for each site. The neutron probe was calibrated by regressing the neutron probe readings against the gravimetric water content measured in soil adjacent to the neutron probe access tubes at three different times, when moisture contents ranged from wet to dry. Volumetric water content was determined as in Gardner (1986) and soil bulk density as in Blake and Hartge (1986). Simpson's rule, a method of numerical integration, was used to calculate the volumetric water content of the 0–80 cm soil layer (Haverkamp et al., 1984). Because plant roots could not establish inside the lysimeters, the water content inside the minilysimeters was usually greater than in the bulk soil outside. To adjust for this, the difference in volumetric water content in the 0–80 cm layer inside and outside the minilysimeters was subtracted from the drainage into the reservoir after each drainage event.

The study sites were monitored twice monthly after the spring thaw from April to December, but only intermittently from January through March, when the ground is often covered with snow. To calculate the annual water balance, monthly precipitation was calculated by using the data from the bulk rain gauges and adjusting against monthly data from long-term weather stations as in Minnich et al. (2000). Monthly standard ET, *sensu* Stephenson (1998) (the term "standard ET" is used here throughout instead of "actual ET") was calculated as in Eagleman (1976). Monthly ET was calculated from the hydrologic equation for April through

December when the sites were visited bimonthly. However, for the periods from January to March, when the sites were only intermittently visited, monthly ET was interpolated by making it proportional to the monthly standard ET for the period in question.

Results

A transition from a strong La Niña in 1988–1989 to a strong El Niño in 1992 occurred during the study period, thus the data include one dry year, one near-normal and one wet year. Precipitation amounts averaged over the three-year period were near the long-term mean for the three sites (Figure 2, Table 1). The annual pattern of precipitation was bimodal, with summer precipitation ranging from 24–30% of annual precipitation. The water balances averaged over three years revealed that ET was 55% of precipitation at the lower forest site, 71% of precipitation at the upper forest site, and 81% of precipitation at the wet-meadow site (Figure 2). Standard ET was 754, 622, and 551 mm, while "climatic deficit" (*sensu* Stephenson, 1998) was 333, 225, and 247 mm for the lower forest, upper forest and wet-meadow sites, respectively.

During the first dry year of 1989–1990, winter (January–March) precipitation was low, but there was significant spring (April–June) precipitation (Figure 3). Winter precipitation was high during the second and third years, but there was little or no spring precipitation. Summer (July–September) precipitation was high during the first and third years, but low during the second year. Evapotranspiration was tightly linked to precipitation at all three sites during the first dry year, but was unusually low and surpluses high during the second year (Figure 3). The low ET during the winter and spring of the second year was likely the result of deep, late snow (Table 2) falling on dry soil. During the wet third year, snow fell early in winter, and both precipitation and ET were higher at the upper forest site.

At the upper forest site, estimated winter ET was near the standard ET, but was considerably higher than standard ET in April (Figure 2). A potential source of error, which would have contributed to the high ET measured at that site, may have been meltwater running off over the frozen ground within the snowpack. The surface soil and lower layers of the snowpack were icy and hard in March of 1991 and 1992 at the upper forest site (Table 2). Snow was observed to remain only a few days at the lower forest site but persisted into April and May at the upper forest site (Table 2).

The lower forest site had low ET in winter and spring, in spite of having nearly double the vegetation cover of the upper forest site (Table 1), apparently because of higher losses to deep drainage at the lower elevation. Significant deep drainage (> 80 cm) was recorded only during the visits of April 1991 and 1992, in response to high winter precipitation. Total deep drainage was 38% and 20% of total precipitation at the lower and upper forest sites, respectively. The higher deep drainage, lower ET, and rapid reduction in available soil moisture at the lower forest site (Figures 2a, 3a) suggest that early melting of snow, followed by infiltration of meltwater into the coarse-textured soil, quickly released surplus water to nearby streams, leaving little moisture for consumption when the vegetation became active later in the spring.

In the mountain meadow, the annual water balance averaged over the study period showed relatively high ET and significant deficits from April through August (Figure 2c). Estimated ET was higher than standard ET during the summer months. Deficits here can be more than made up by both run-in and interflow from the surrounding uplands. The water balance averaged over three years revealed that average summer precipitation was slightly greater than ET at both forest sites (Figure 2). Although peak summer ET was about the same at both sites, summer ET was generally higher at the warmer lower forest than at the upper forest site

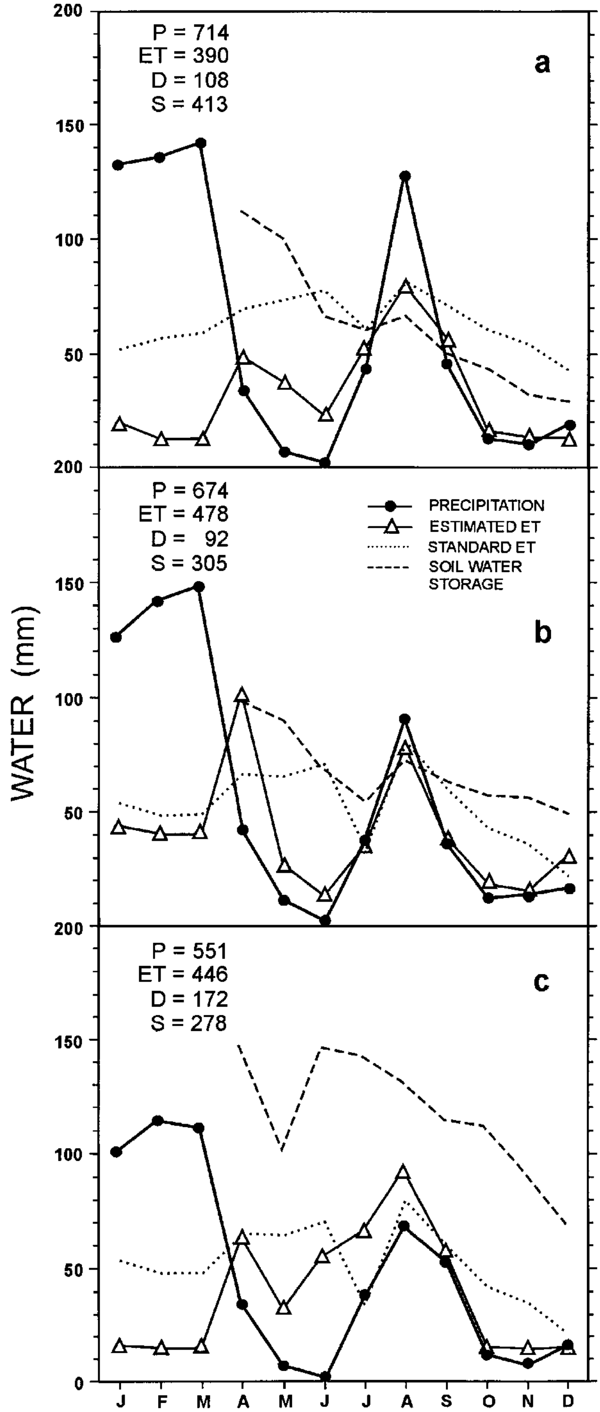

FIGURE 2 Annual water balance (0–80 cm soil layer) averaged over the three-year study period at the (a) lower forest, (b) upper forest, and (c) wet-meadow sites. P = precipitation; ET = estimated evapotranspiration; D = apparent deficit (sum of ET – P for the months when ET > P); and S = apparent surplus (sum of P – ET for the months when P > ET).

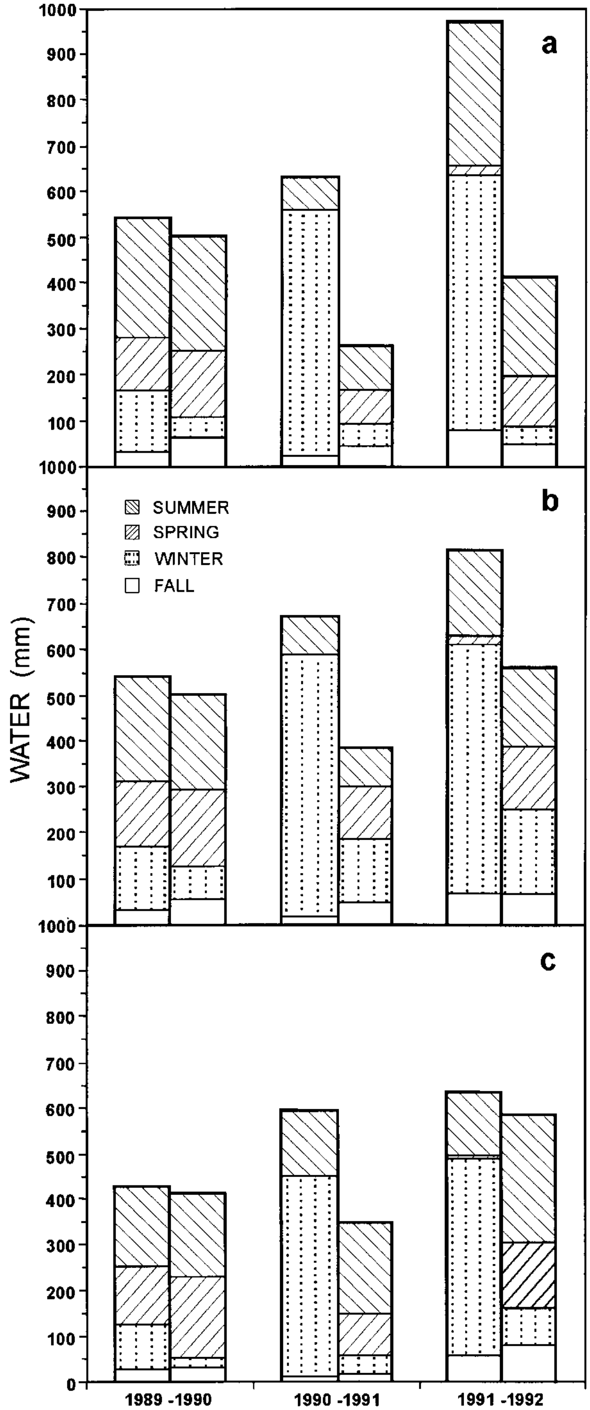

FIGURE 3 Annual precipitation (left column) and ET (right column) and their seasonal contributions at the (a) lower forest, (b) upper forest, and (c) wet meadow sites.

TABLE 2 Snowpack Measurements at Study Sites in the Sierra San Pedro Mártir (Values are means ± 1 standard deviation.)

Site	Date	Snow depth	Water content cm	Density %	*n*
Mt. Meadow	10 March '91	35.3 ± 8.4	13.1 ± 3.1	37.2 ± 0.0	8
	7 January '92	20.0 ± 3.2	7.4 ± 1.2	37.2 ± 0.0	8
	12 March '92	31.8 ± 4.9	11.7 ± 1.8	37.3 ± 0.1	8
	22 December '92	23.5 ± 1.4	8.7 ± 5.3	37.2 ± 0.0	5
Upper Forest	10 March '91	49.4 ± 3.4[a]	34.9 ± 3.0	70.8 ± 6.6	8
	7 January '92	27.3 ± 6.5	8.1 ± 2.7	30.0 ± 7.3	8
	12 March '92	91.8 ± 12.2[bc]	65.2[bc]	—	12
	22 December '92	38.5 ± 1.6	13.2 ± 1.4	34.4 ± 4.8	5
Lower Forest	22 December '92	17.2 ± 0.5	6.4[d]		

[a]Lower snowpack icy, great difficulty boring down to soil.
[b]Unable to bore down to soil, so underestimated snowpack depth and water content.
[c]Estimated by using density of 71%.
[d]Estimated by using density of 37%.

(Figures 2, 3). High summer precipitation at the lower forest site in 1992 (Figure 3a) skewed the average, resulting in a net summer surplus (Figure 2a), likely because intense summer rainfall induces proportionately higher runoff and lower soil intake. Indeed, during this period, soil water storage at the lower forest site registered an increase of only 42 mm in response to 222 mm of precipitation. Summer ET, however, appears to be driven mostly by summer precipitation.

The relationship between summer precipitation and ET was strong at all three sites (Figure 4), except that at the mountain meadow there was much higher summer ET after the wet winter of 1992; this was likely supplied from stored soil water and interflow. The outlier made the relationship between precipitation and ET meaningless for the wet meadow, indicating that after a wet winter, summer ET at the meadow is independent of summer rainfall. The regression implies that 50 mm of water are mobilized in summer from water stored during winter, and that 63% of summer precipitation is evapotranspired during the same summer. This apparently low rate of usage of summer precipitation was likely due to the upward skewing of average summer precipitation by a single event in the summer of 1992.

Discussion

The relationship between average precipitation and ET for the three study sites in the Sierra San Pedro Mártir was similar to published long-term ET data for similar sites having extreme water deficits (following the classification of Rutter, 1968) in southern California and Israel, as well as for ET (calculated as precipitation minus streamflow) for 18 watersheds in California south of the 35th parallel (Figure 5). Our results are also consistent with a runoff coefficient of 19% (that is, ET of 81%) calculated for the period 1963–1983 by Orozco-Zavala (1991) for the entire Rio Santo Domingo watershed, which includes the lower forest site. Average losses by ET were 69% of precipitation in the present study. The difference between our results and those of Orozco-Zavala (1991) may be, (1) that the Rio Santo Domingo watershed, which covers 1,227 km^2, includes extensive low-elevation shrublands that are more arid than the forested uplands of the Sierra San Pedro Mártir, and (2) that plant roots can extract deep water below the solum (Arkley, 1981; Sternberg et al., 1996; Hubbert, 1999).

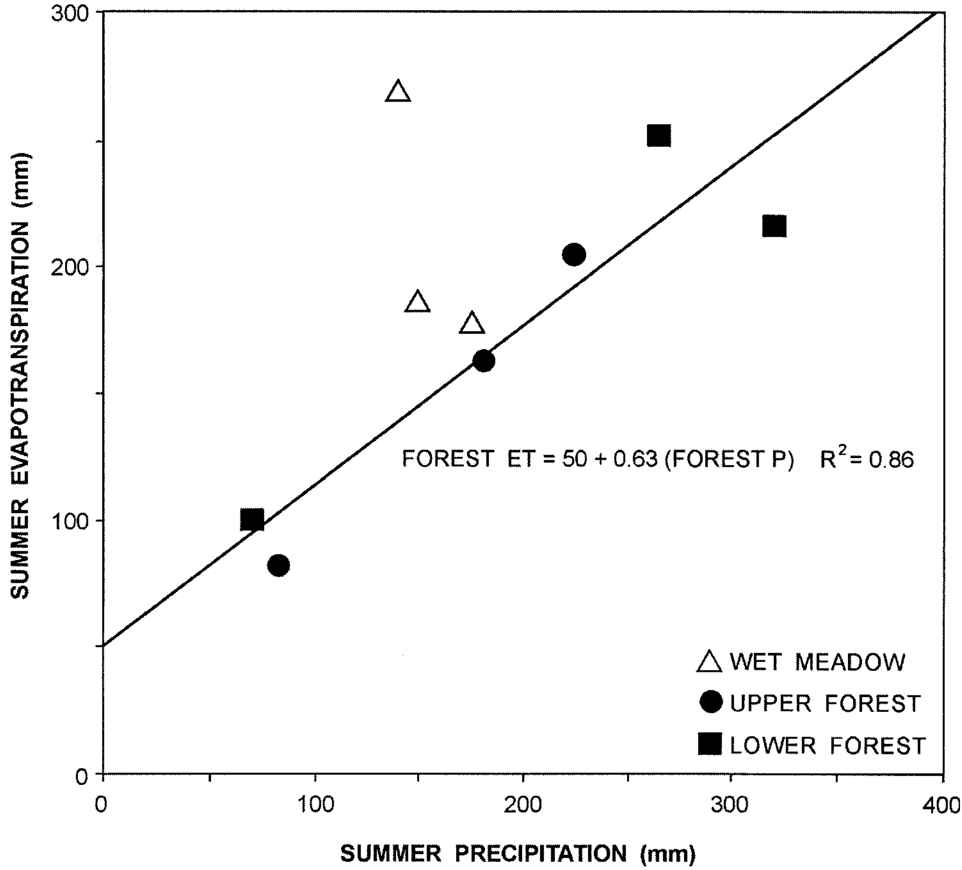

FIGURE 4 Relationships between summer precipitation and summer ET.

The water budgets in this study show, counterintuitively, that the lower elevation forest had lower ET in winter than the forest at the higher elevation (Figures 2, 3). Warmer temperatures at lower elevations favor earlier melting of snow, resulting in higher losses to deep drainage and streamflow, and leaving little moisture for consumption later in spring. The persistence of the snowpack at higher elevations contributes to higher ET in winter and spring by reducing deep drainage and postponing infiltration into the soil until late spring. Rapid and early melting of the snowpack at the lower elevations would also limit losses from meltwater running off over the frozen ground. Although the relationship between summer precipitation and summer ET at the forest sites was linear (Figure 4), that between precipitation and ET in the cold months was meaningless. This was likely due both to high variability in winter precipitation and the differences in ET and deep drainage between the lower and upper forest sites noted above.

Analysis of Data in Review Studies

Regression analysis of worldwide data published in review studies of both streamflow and evapotranspiration showed a consistent increase in the strength of the relationship between precipitation and ET with increasing water deficit, and a concomitant decrease in the strength of the relationship between precipitation and streamflow (Table 3). It seems reasonable that under extreme water deficits, when

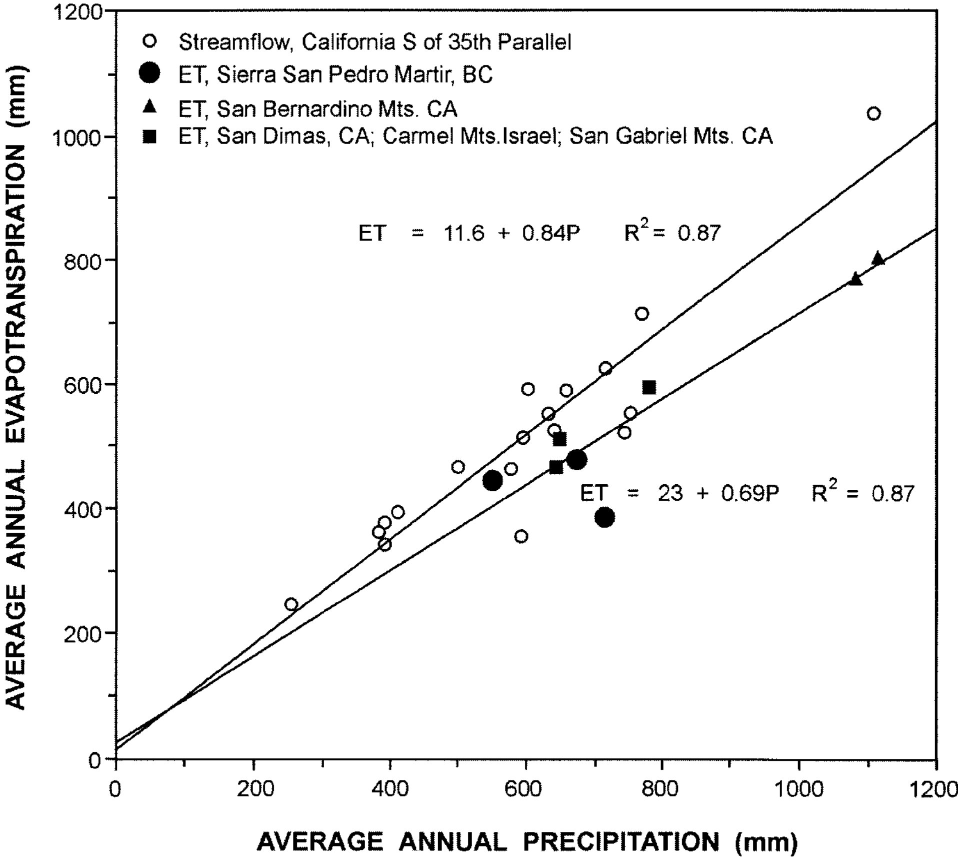

FIGURE 5 Relationships between average precipitation and ET for available data from extreme-deficit watersheds in streamflow and ET studies. Evapotranspiration was calculated as precipitation minus streamflow for the streamflow studies. Streamflow data are from Turner (1986). Evapotranspiration data for San Dimas and the Carmel Mountains and San Gabriel Mountains are from Rutter (1968), and data for the San Bernardino Mountains are from Arkley (1981). Data for San Dimas and the San Bernardino Mountains were modified as explained in the Appendix.

losses by ET are 70–80% of precipitation, the ET regression line should explain most of the variability in the data. For the combined worldwide data in streamflow studies (n = 168), 76% of precipitation was lost through streamflow and 24% through ET, while for ET studies (n = 37) the proportions are 68% and 32%. Classifying on the basis of water deficits, along a continuum from watersheds, where water loss is dominated by streamflow, to those where it is dominated by ET, reduces the noisiness of the data and provides physical explanations for the slope and intercept of the regression line.

The slope of the regression line indicates the portion of precipitation that is lost to streamflow or ET. That the sum of the slopes of the regression lines equals one (Table 3) indicates how precipitation is partitioned between the two processes. The explanation for the intercept appears to be that, in most systems, the soil profile must be recharged before ET can take place. Conversely, the negative intercept for the streamflow regressions indicates the amount of precipitation that must first go to

TABLE 3 Regression Analyses of Average Annual Precipitation (P) vs. Average Annual Streamflow (S, calculated as P-ET), and Average Annual Precipitation vs. Average Annual Evapotranspiration (ET, calculated as P-streamflow) from Published Streamflow and Evapotranspiration Studies

Relationship	Intercept	Slope	R^2	*P* value	*n*	Water deficit category	Location	Reference
Streamflow Studies								
P vs. ET	504	0.16	0.20	<0.0001	78	Negligible to extreme	Worldwide	Bosch & Hewlett 1982
P vs. S	−504	0.84	0.88	<0.0001	”	”	”	”
P vs. ET	311	0.20	0.46	0.0005	22	Small to extreme	USA, Europe	Shachori & Michaeli 1965
P vs. S	−311	0.80	0.93	0.0005	”	”	”	”
P vs. ET	210	0.45	0.76	<0.0001	68	Small to extreme	California	Turner 1986
P vs. S	−210	0.55	0.82	<0.0001	”	”	”	”
P vs. ET	11.6	0.84	0.87	<0.0001	18	Severe to extreme	California S of 35th Parallel	Turner 1986
P vs. S	−11.6	0.16	0.19	0.07	”	”	”	”
Evapotranspiration Studies								
P vs. ET	345	0.19	0.34	<0.05	12	Negligible to extreme	Worldwide	Rutter 1968
P vs. S	−345	0.81	0.91	<0.0001	”	”	”	”
P vs. ET	86	0.43	0.71	<0.0001	17	Small to severe	Worldwide	Rutter 1968[a]
P vs. S	−86	0.57	0.58	0.0004	”	”	”	”
P vs. ET	23.6	0.69	0.87	0.0008	8	Extreme	Californias, Israel	Rutter 1968[b], Arkley 1981[c], this study
P vs. S	−23.6	0.31	0.58	0.03	”	”	”	”

Data for North Fork[a], San Dimas[b], and San Bernardino Mountains[c], California, were modified as explained in the Appendix.

recharge the soil profile before streamflow can begin. Interestingly, the regressions for streamflow data for California south of the 35th parallel (Turner, 1986), which include a number of desert and low-elevation watersheds, as well as the available ET data for extreme-deficit environments, both have near-zero intercepts (Table 3, Figure 5). The near-zero intercept indicates that in watersheds under extreme water deficits—mostly in low latitudes and elevations—any precipitation that falls is almost immediately available for evaporation. On the other hand, watersheds in which deficits are low and streamflow predominates—mostly in higher latitudes and elevations—precipitation must first recharge the soil profile (often as thawing snow) before streamflow can occur. The zero-intercept is consistent, therefore, with watersheds that receive sparse and mostly liquid precipitation.

The difference in the slopes of the regression lines between ET studies (0.69) and streamflow studies (0.84) in Figure 5 may represent extraction of water by deep roots, that is, deeper than the layer of soil usually considered in ET studies. This loss of water from deep cracks and decomposed rock has been difficult to quantify, but comparison of the results of streamflow and ET studies, especially in regions of severe to extreme water deficits, may be one way to constrain estimates of this important process. It must be noted that the slope of the ET regression line in Figure 5 depends greatly on the values at the high end of the precipitation range. Arkley (1981) estimated ET at study sites in the San Bernardino Mountains during the cold winter months (November to April) by using potential ET (PET) calculated by the Thornthwaite method. This likely underestimated annual ET, and resulted in an unreasonably low slope of 0.39 when ET was regressed against precipitation. To correct this underestimation, standard ET was calculated for the cold months as in Eagleman (1976) and substituted for PET in Arkley's data (see Appendix). Additional data are required for locations receiving between 800 and 1000 mm of precipitation to improve the estimate of the slope of the ET regression line.

Conclusions

The results of this study demonstrate that ET at the southern limit of the chaparral-forest boundary in Baja California is comparable to that of sites having similar precipitation and vegetation regimes in neighboring southern California. The results of streamflow and soil ET studies are consistent with each other, and show that classifying by the degree of water deficit reduces the noisiness of the data. Considering only data from regions of extreme soil water deficit results in high correlations between precipitation and ET in both streamflow and evaporation studies. Moreover, the intercept and slope of the regression lines provide information on soil water recharge and the proportion of precipitation that is lost through either ET or streamflow. Comparing the results of streamflow and ET studies in watersheds under extreme water deficits suggests that about 70% of precipitation is lost by ET from the soil, while the remaining 30% is about evenly partitioned between extraction of water by deep roots below the solum and streamflow.

Appendix

1. Only quantitative data were used from the review studies cited in Table 3; approximations were not considered. To calculate average precipitation, ET, and surplus for the longest terms possible, original sources were consulted and data were manipulated as follows: (a) Data for the "undisturbed" plot at North Fork, California, in Table 9 of Rowe (1948) were combined with that for the "natural" plot in Table 4 of Rowe and Colman (1951) to give average precipitation, ET, and surplus for the period 1934–1940 of 1,043, 466, and 577 mm, respectively, and (b) data for "five dry years" and one "wet" year at the San Dimas large lysimeters (Patric, 1961)

were combined to give average precipitation, ET, and surplus for the period 1952–1958 of 642, 466, and 176 mm, respectively.

2. Water balance data in Table 6 of Arkley (1981) were modified by substituting standard ET, calculated as in Eagleman (1976), for Arkley's calculated potential ET for the cold months (November to April), by using long-term weather data for Fawnskin (2,088 m) in the San Bernardino Mountains (USDA Forest Service, 1982). A correction factor was developed which gave precipitation, ET, and surplus of 1,114, 804, and 310 mm for the Dogwood 2, and 1,082, 769, and 313 mm for the Breezy Point sites, respectively.

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